Numerical modeling of the response of alluvial rivers to Quaternary climate change

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Abstract

A numerical model, which simulates the dynamics of alluvial river channels on geological (Quaternary) time scales, is presented. The model includes water flow, channel dimensions, sediment transport and channel planform type. A number of numerical experiments, which investigate the response of an alluvial river to imposed sequences of water and sediment supply, with special emphasis on the time lags between these controlling variables, as well as a downstream discharge increase, are presented. It is found that the influence of the time lags can be substantial, having major implications for the reconstructions of palaeo climate based on river channel behavior documented in the geological record. The model is further applied to both a conceptual warm–cold–warm cycle and a reconstructed evolution of the river Meuse, the Netherlands, during the Late Glacial–Holocene warming. Results show that the model is capable of explaining the response of this river, although better validation against palaeoenvironmental data remains necessary. © 2000 Elsevier Science B.V. All rights reserved.

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1. Introduction

During the last decades, substantial progress has been made on the understanding of the evolution of large river systems during the Quaternary, especially as a function of the major climatic changes. For examples based on the Rhine–Meuse delta, see for instance (Vandenberghe, 1993a, 1994; Kasse et al., 1995; Huisink, 1997, 1998; Törnqvist, 1998; Tebbens et al., 1999). Despite these research efforts, little is known about this evolution from a process-based viewpoint. This is due to a number of reasons: (1) the inherent complexity of the fluvial system (Hey, 1979a,b; Phillips, 1991); (2) the failure to date fluvial sediments accurately enough to link them with palaeoenvironmental change on the scale of hundreds of years (Lowe and Walker, 2000; Kasse et al., 2000); and (3) the previous emphasis on reconstruction of the palaeoenvironment based on geomorphology (terrace mapping), sedimentology (facies reconstruction), and palaeoecology (vegetation reconstruction), where geomorphological processes have been inferred from the traces they left in the landscape.

The aim of this paper is to enhance our understanding of the way fluvial systems respond to environmental change in general, and warm–cold (such as glacial–interglacial) transitions in particular. The approach adopted here is to explore the prospects of...
numerical modeling of fluvial geomorphological processes in order to increase our understanding of the fluvial landscape evolution. Development of this model is driven both by field observations and by state of the art geomorphological process knowledge.

2. Theory

Our model aims at the simulation of river morphodynamics as a function of climate-related boundary conditions like water discharge and sediment supply. The model is implemented as a system subdivided into three main modules: channel morphology, water flow and sediment transport. The cyclic interdependence of these modules, as depicted in Fig. 1, reflects the mutual dependency of form and process within the fluvial system. A negative feedback between the sediment transport and channel morphology causes the system to be inherently stable (that is, on macro scale, see Phillips, 1990 for a discussion on inherent stability on smaller scales), and a dynamic equilibrium could be reached. The interplay between the system’s relaxation time and the frequency of external events will determine if this equilibrium will be reached eventually. For example, if climate changes faster than the time span needed for adaptation by the river system, equilibrium will not be reached. If changes occur less frequently, equilibrium will be reached, although for a limited duration (Bull, 1991). The status of the fluvial system’s components as either ‘dependent’ or ‘independent’ of other components is a matter of scale (Schumm and Lichty, 1965; Lane and Richards, 1997). For instance, on short time scales, channel slope is independent of sediment transport, but on (much) larger time scales it is dependent, because channel slope slowly adjusts to the prevailing sediment transport regime by means of incision/aggradation and/or sinuosity adjustment. Within the context of this study, we regard all the system’s components as dependent of each other.

In the next sections, the model’s main modules will be presented with respect to their physics and numerical modeling.

2.1. Channel morphology

The state of the river morphology is described in terms of riverbed elevation $z$ [L], bankfull channel width $B$ [L], bankfull channel depth $y$ [L], and derivative variables like channel slope $S = \frac{\partial z}{\partial x}$ [L/L].

The most important processes that act on the river morphology are incision and aggradation, which are actually the change of riverbed elevation over time, $\frac{\partial z}{\partial t}$, [L T$^{-1}$]. This erosion/sedimentation rate is directly governed by the spatial divergence of sediment transport, as can be seen from the one-dimensional continuity equation:

$$\frac{\partial z}{\partial t} = \nabla \cdot q_s + I + T = \frac{\partial q_s}{\partial x} + I + T$$

where $q_s$ [L$^2$ T$^{-1}$] is volumetric unit sediment transport, $I$ [L T$^{-1}$] is unit external sediment input (e.g. from banks, tributaries), and $T$ [L T$^{-1}$] is vertical movement due to tectonics. For simplicity’s sake, $I$ and $T$ are ignored in the current experiments.

Channel width $B$ [L] is one of the least understood morphological variables. Although some progress is made in process-based prediction of channel width (Ikeda et al., 1988; Ikeda and Izumi, 1990, 1991), it is generally calculated by a simple power–law relationship of the form:

$$B = aQ^b$$

where $Q$ [L$^3$ T$^{-1}$] is usually the bankfull discharge and exponent $b$ is 0.5. Within the model, the relation:

$$B = 4Q^{0.5}$$

is being used (van den Berg, 1995).
In our model, mean channel depth \( y \) [L] is assumed to be equal to the average flow depth under bankfull conditions, an assumption that can be justified on base of the dependency of channel cross-sectional shape on bankfull discharge.

2.2. Water flow

Water flow parameters like shear stress need to be included in the morphodynamic model because these directly determine sediment transport. The relevant flow variables are depth-averaged flow velocity \( U \) [L T\(^{-1}\)], and bottom shear stress \( \tau \) [M L\(^{-1}\) T\(^{-2}\)], while the relevant flow parameters are kinematic viscosity of water \( \nu \) [L\(^2\) T\(^{-1}\)], mass density of water \( \rho \) [M L\(^{-3}\)] and flow resistance \( n \) (Manning’s \( n \) ) [T\(^{-1/3}\)].

Flow through open channels can be modeled in several degrees of complexity. For a complete physically correct simulation of the three-dimensional temporal variable flow field within the river channel, the Navier–Stokes equation is required. This equation is, however, non-linear, and therefore time-consuming to solve. Simplifications can be made by relaxing some of the constraints for this equation, such as the well-known kinematic wave approximation (Dingman, 1984). The simplest model is that of steady, uniform flow. Although not applicable to short-term situations where flow conditions vary significantly, this model can be assumed to perform well over the time spans that we are interested in, since individual events are ignored and overall flow conditions vary slowly (Howard, 1994).

Calculation of flow is based on the combination of continuity of water and the Manning’s equation for flow resistance:

\[
Q = BUy
\]

\[
U = \frac{1}{n} R^{2/3} S^{1/2}
\]

where \( Q \) is the (bankfull) discharge, \( B \) is the mean channel width, \( R \) is the hydraulic radius \( U \) is depth-averaged flow velocity, \( S \) is channel slope, and \( n \) is Manning’s \( n \).

If we assume, for simplicity’s sake, that for wide, shallow (alluvial) channels:

\[
R \approx y,
\]

then, without loosing much precision, we can solve Eqs. (4) and (5) for \( R \) as:

\[
R = \left( \frac{Qn}{B\nu S} \right)^{0.6}
\]

\[
U = \frac{Q}{BR}.
\]

Bottom shear stress \( \tau \) is then calculated by:

\[
\tau = \rho gSR.
\]

2.3. Sediment properties and transport

Sediment in general is being described by its size distribution, density and shape. Within our simplified model, only density \( \rho_i \) [M L\(^{-3}\)], and median size \( D_{50} \) [L] are being considered. Sediment properties are being held constant over time in the current experiments.

Sediment transport is still one of the least understood processes in fluvial hydraulics. This is mostly due to the inherent complexity of the factors involved: natural water flow is characterized by chaotic movements on several spatial and temporal scales (turbulence), while natural sediment consists of a mixture of grain sizes and natural channels exhibit a complex morphology made up of banks, bars and pools.

Despite this complexity, several sediment transport equations have been formulated, mainly based on flume experiments. From the numerous equations known from literature, only one has been selected: the Einstein–Brown bedload transport equation. This model was selected because of its successful use in previous modeling attempts on geological time scale (Howard, 1982, 1994; Howard et al., 1994; Willgoose et al., 1991).

2.3.1. The Einstein–Brown transport equation

The Einstein–Brown sediment transport equation (Brown, 1950) is an empirical relationship between non-dimensional sediment transport \( \Phi \) and non-dimensional water flow \( 1/\Psi \). \( \Phi \) is defined as:

\[
\Phi = \frac{q_s}{\rho_i F \left( g (s-1) d_i^3 \right)^{1/2}}
\]
where \( q_s \) is sediment transport per unit width [M L^{-1} T^{-1}], \( g \) is the gravitational acceleration, \( s \) is the specific sediment density \( \rho / \rho_c \), \( d_s \) is the median sediment size, and the parameter \( F \) [-] is a measure for sediment flow velocity and is defined as:

\[
F = \sqrt{\frac{2}{3} + \frac{36 \nu^2}{gd_s^2 (s - 1)}} - \sqrt{\frac{36 \nu^2}{gd_s^2 (s - 1)}}. \tag{11}
\]

The non-dimensional flow number \( 1/\Psi \) is defined as:

\[
\frac{1}{\Psi} = \frac{\tau}{\rho g (s - 1)d_s^2}. \tag{12}
\]

Transport and flow numbers are coupled by the empirical relation:

\[
\Phi = 40 \left( \frac{1}{\Psi} \right)^{3}. \tag{13}
\]

Assuming homogeneous sediment, these equations, when combined with the uniform flow model, can be simplified to (Willgoose et al., 1991):

\[
q_s = F_2 (RS)^{3} \tag{14}
\]

\[
F_2 = 40 \rho_e F \sqrt{\frac{g (s - 1) d_s^2}{(s - 1)d_s}} \left[ \frac{1}{(s - 1)d_s^2} \right]^3. \tag{15}
\]

Assuming a wide channel with uniform depth, and ignoring the (small) error introduced by setting \( R = y \), and using Manning’s equation, we obtain:

\[
q_s = F_2 n^{1.8} q^{1.8} S^{2.1} \tag{16}
\]

where \( q \) is specific water discharge \( Q/B \). Eq. (16) is of the same power-law form like used in several landscape evolution models (e.g. Willgoose et al., 1991; Howard, 1994; Howard et al., 1994).

Since \( q_s \) is in units of [M T^{-1} L^{-1}], a volumetric sediment transport rate \( q_{v_s} \) [L^{2} T^{-1}] is calculated as:

\[
q_{v_s} = \frac{q_s}{\rho_{v_s}} \tag{17}
\]

where \( \rho_{v_s} \) is volumetric bulk density of the sediment. This volumetric sediment discharge can be inserted into Eq. (1).

### 2.4. Tectonics

The influence of tectonically induced vertical movement on the river profile can be incorporated straightforward in the model by including a spatial heterogeneous uplift rate \( T \) that acts as a boundary condition for the model (in Eq. (1)). Processes like (local) uplift and/or subsidence (Snow and Slingerland, 1987, 1990) or sea (base)-level change (Bonneau and Snow, 1992) can be simulated in this way. Although (neotectonic) tectonics are known to have affected the Meuse River (Houtgast and van Balen, 2000; Stouthamer and Berendsen, in press), this paper focuses on climate-induced river response. Therefore, this option is not used within the experiments described here.

### 2.5. Time scales

Within (Quaternary) geology, time scales are extremely important, since climate parameters like precipitation and temperature are highly dynamic on time scales from hours up to >100 ka (Vandenberghe, 1995).

- On the scale of hours to days, individual events like summer thunderstorms are important.
- Within 1 year, the weather changes seasonally. Example: snow melt in spring.
- On the decadal scale: gradual climate change. Example: global warming.
- On the centennial to millennial scale, we can distinguish climatic events like the Younger Dryas (Isarin et al., 1998) or the Heinrich Events (Heinrich, 1988; Bond et al., 1992).
- On the 10-ka time scale, transitions between periglacial and more temperate periods appear to be important for fluvial evolution, which is determined by the non-linear response of climate derivatives as vegetation and soil cohesion (Vandenberghe, 1993a). Example: the Weichselian Pleniglacial–Late-glacial transition.
- On the 100-ka time scale, glacial and interglacial can be distinguished. Example: Weichselian, Holocene. Fluvial evolution is climatically dependent.

Although the combination of events on all time scales together shape the landscape, it is a general assumption within geomorphology that the events...
that have median magnitude and recurrence interval have the greatest influence on long-term landscape evolution (Wolman and Miller, 1960). Although this concept is currently under attack (Richards, 1999), it can still be used as a valuable assumption that enables the use of ‘effective’ discharges to be used as an approximation for the natural discharge variation.

A proper modeling effort should include all the different time scales, which in fact means that the time scale of the model should be adapted to the shortest time scale involved, which is hours. This is possible by using hydraulic engineering and hydrological routing models that operate on this time scale, using time steps and time-dependent input of hourly resolution. It is, however, not feasible to use these models for the longer time scales (> 1000 year), which are the subject of this paper. The short-scale dynamics are therefore lumped into a ‘channel-forming’ discharge, which is usually taken to be the bankfull discharge, and the time scale of modeling becomes years, which is more appropriate for experiments concerning Quaternary climate change.

2.6. Channel pattern

Channel pattern, the planform state of alluvial rivers, is generally considered to be a continuum, ranging from straight via meandering to finally braided rivers (Leopold and Wolman, 1957; Ferguson, 1987; Alabyan and Chalov, 1998). However, a problem remains with classifying single-channel/anastomosing rivers. There is, therefore, a growing feeling that proper pattern classification is more complex and should consider multiple independent controls, where, for example, braided/meandering is independent from single channel vs. anastomosing (Schumm, 1985; Bridge, 1993).

Within this study, we use a method based on van den Berg (1995), who was able to successfully separate braided from meandering rivers, based on two independent variables: potential unit streampower $\omega_p$ [W/m²] and median grain size $D_{50}$ [m] (see Fig. 2). In our model, the ratio of actual potential unit streampower (calculated from the model state) to threshold streampower (the division line in Fig. 2) is used as a ‘pattern index’. Values $> 1$ indicate braiding and values $< 1$ indicate meandering. Given his equation for potential unit streampower under bankfull conditions ($Q = Q_{bf}$):}

$$\omega_p = \begin{cases} 2085S\sqrt{Q_{bf}} & \text{for sand bed rivers} \\ 3266S\sqrt{Q_{bf}} & \text{for gravel bed rivers} \end{cases}$$

and for threshold stream power:

$$\omega_c = 900D_{50}^{0.42}$$

it follows that braiding occurs when:

$$S\sqrt{Q_{bf}}D_{50}^{-0.42} > 0.432$$

for sand bed rivers, and:

$$S\sqrt{Q_{bf}}D_{50}^{-0.42} > 0.276$$

for gravel bed rivers.

It must be emphasized, however, that some nonlinear effects during the transition from braided to meandering streams might be expected, for instance due to differences in width–depth ratio for the different regimes. These effects are not included in the model.

2.7. Climate change

Climate must be quantified using at least two different variables—temperature and precipita-
tion—and two numbers per variable: a measure of the absolute amount, like annual mean, and a measure of the within-year variability, like seasonality (Bull, 1991). Seasonal effects are currently ignored in our model; precipitation and temperature are described using a single number, which acts as an ‘effective’ value. Alluvial rivers do not respond to climate directly. Since a river can be viewed as a concentrated flow of water and sediment through the landscape, it is the input of water and sediment that drives the river system. River and climate are therefore indirectly coupled.

Water flowing in river channels is derived from several sources: rainfall directly into the channels, groundwater seepage, and (sub)surface runoff from hillslopes. Since the areal extent of alluvial channels is small compared to their basins, their direct input can be neglected. Under steady state conditions, seepage, subsurface and surface runoff are all considered to be in equilibrium with precipitation, which is a reasonable assumption on the proper centennial time scales.

However, not all rainfall is converted to runoff. A significant fraction of the rainfall returns to the atmosphere by evapotranspiration. The amount of evapotranspiration is a function of vegetation physiology and cover, temperature, windspeed and soil physical properties. Within the model, these processes are ignored and bankfull discharge is the direct driving force, representing the rainfall ending up in the river.

Sediment supply is, like water supply, to a great extent regulated by vegetation (Bull, 1991; Leeder et al., 1998). In our model, sediment supply to the river channel is equal to the sediment yield from hillslopes. Sediment yield is a function of erosivity and erodibility (Morgan, 1986). Erosivity is linked to the (temporal distribution of) precipitation, while erodibility of unconsolidated sediments is linked to the protection against soil erosion by a vegetation cover. These processes are not modeled explicitly, and sediment production is taken as a climate-dependent parameter.

To summarize, the process cascade can be outlined as:

\[
\begin{align*}
\text{precipitation} &= F(\text{climate}) \\
\text{temperature} &= F(\text{climate}) \\
\text{vegetation} &= F(\text{precipitation}^+, \text{temperature}^+) \\
\text{evapotranspiration} &= F(\text{vegetation}^+, \text{temperature}^+) \\
\text{runoff} &= F(\text{precipitation}^+, \text{evapotranspiration}^-) \\
\text{sediment yield} &= F(\text{runoff}^+, \text{vegetation}^-) \\
\text{discharge} &= F(\text{runoff}^+) 
\end{align*}
\]

which can be summarized into:

\[
\begin{align*}
\text{sediment yield} &= F(\text{climate}) \\
\text{discharge} &= F(\text{climate})
\end{align*}
\]

This later simplified scheme is used in our model, that is, sediment input and discharge input are directly specified as boundary conditions.

2.8. Implementation

The equations describing the fluvial processes are solved using a one-dimensional, explicit finite difference method. This approach enforces the computational time step to be very small (days), in order to maintain numerical stability. The model uses a computational grid, which has a total span of 100 km and a resolution of 4 km. This resolution is an acceptable compromise between spatial detail and computational efficiency. The model results show no sensitivity to this numerical discretization. In the next section, results are shown for the computational node located near the upstream discretization. During each simulation step, the following sequence of calculations has been used:

1. input discharge \( Q \),
2. calculate channel gradient \( S \) from \( \frac{\partial z}{\partial x} \),
3. calculate channel width \( B \) from Eq. (3),
4. calculate hydraulic radius \( R \) from Eq. (7),
5. calculate unit sediment transport rate \( q_s \) from Eqs. (10)–(17),
6. calculate spatial divergence of sediment transport rate \( \frac{\partial q_s}{\partial x} \), using external sediment input as the upstream boundary condition,
7. calculate change in elevation \( \frac{\partial z}{\partial t} \) from Eq. (1),
8. update elevations \( z_{t+\Delta t} \).
3. Numerical experiments

In the next few sections, a number of modeling experiments are presented. In experiment 1, the influence of individual changes in discharge and sediment supply on river evolution is investigated. In the second experiment, discharge and sediment supply are varied both, with different time lags between them. In experiment 3, a downstream increase in water discharge is applied to study the response of the river to this spatial pattern. A hypothetical warm–cold–warm scenario is applied in the fourth experiment. Finally, the evolution of the Meuse river (the Netherlands) during the Weichselian–Holocene transition is being simulated in the fifth experiment.

3.1. Changes in bankfull discharge and sediment supply

In this first experiment, the bankfull discharge and sediment supply are changed in a step-wise manner. Simulation results are shown in Fig. 3. The river reacts on increased discharge by incision. The reason for this behavior is that an increased discharge causes the transport capacity to increase also, while sediment supply stays at the same level. The river scours its bed until a new equilibrium state is reached in which discharge and sediment supply are in equilibrium, which is realized by decreasing the longitudinal gradient.

A decrease in discharge leads to aggradation, since the reduced streampower is no longer able to transport all the sediment supplied. Note that the amount of increase and decrease in sediment is both 50% of the ‘default’ values, but that the response in terms of meters incision/aggradation is asymmetric. This can be explained from the non-linear relation between sediment transport and discharge and slope.

When sediment supply increases while the transporting capacity remains the same, less sediment is transported out of the channel, leading to aggradation within the channel until a new steeper equilibrium profile is reached. A similar decrease in sediment causes an incision. Again, changes are again 50% of default values, but in this case, the response is more symmetric.

The response of the ‘channel pattern index’ is shown in Fig. 3. The most important feature are

![Fig. 3. Results of experiment 1: stepwise variations in bankfull discharge and sediment supply. The top two panels are model input, the bottom panels are output. The dashed line in panel (D) indicates the braided/meandering threshold.](image_url)
again asymmetries in between changes in pattern index and changes in the forcing factors, although less than elevation. In this particular example, only the ‘increased sediment supply’ scenario is able to change the system towards braiding, which does not necessarily mean that discharge is not able to do so. Comparing the response of elevation and pattern index leads to the conclusion that given equal changes in the forcing factors (±50%), changes in discharge have the most impact on incision/aggradation, and changes in sediment supply have the highest impact on channel pattern.

3.2. The impact of time lags

In experiment 2, bankfull discharge and sediment supply are varied simultaneously. However, for four different runs, the time lag between maximum discharge and maximum sediment supply varies between no lag (in phase) and 5000 year (out of phase). As can be seen in the results depicted in Fig. 4, the responses of the river system to these time lags differ greatly.

If discharge and sediment supply are perfectly in phase (solid lines), then the vertical adjustments of the river channel are relative small. The reason for this is that the increased sediment supply is compensated by an increase in transport capacity. Hence, there is no net incision or aggradation. The sediment just flows faster through the channels.

If discharge and sediment supply are in antiphase (dash-dotted lines), then low transport capacities are combined with high sediment supplies and, as a result, aggradation is maximal, while high transport capacity and low sediment supply lead to maximal incision. For time lags that are between these extremes, the response is still quite large, indicating that, within this experiment, the response is not proportional to the time lag.

The response of channel pattern index (panel D) indicates that pattern switching is less sensitive for the imposed time lags. The timing of braiding phases correlates with sediment supply maxima, again indicating that sediment supply has a larger impact on channel pattern than discharge. The no-lag case (solid lines) indicates that aggradation/incision can be independent of channel pattern change. Here, a phase of incision is associated with a braiding pattern.

While this seems illogical at first sight, it is a consequence of the dynamics of the system when...
viewed as controlled by multiple conditions (sediment and water supply).

3.3. The impact of downstream changes in discharge

In the third experiment, bankfull discharge (Fig. 5C) is varied in space and time. The spatial variation is a gradual downstream increase in discharge, where we assume that discharge is proportional to drainage area, which itself increases quadratic with downstream distance. For simplicity’s sake, discharge is kept constant with time. Upstream sediment supply (Fig. 5B) is changing with time, modelled as a sinuous wave with a period of 10,000 year. The resulting longitudinal profiles and channel patterns are shown in Fig. 5D and A.

Three reaches can be distinguished in Fig. 5A. In region 1, the most upstream reach, the river channels are always of the braided type. In region 3, the most downstream reach, the river is constantly in a meandering state. In region 2, the middle reach, the river behavior is more dynamic. Here, the channel pattern threshold is crossed several times, indicating that the river switches between the two states, from meandering to braided and back.

These results correspond with the geomorphologists’ common knowledge that braided channels are more likely to be found in the upper stream reaches and meandering channels in the more downstream reaches. Field observations that are not compatible with this scheme require careful examination, and alternative scenarios for downstream or temporal variations of water and sediment supply must be formulated.

3.4. A conceptualized warm–cold–warm transition

In experiment 4, we study how a model river might react on a scenario that represents, in a simplified way, the warm–cold–warm transitions that are representative for the Quaternary (Fig. 6a). This transition is formalized as an input succession of water and sediment supply, which has been adapted from Vandenberghe (1993a).

Fig. 5. Results of experiment 3: downstream and temporal variability of channel pattern. (A) Spatio-temporal evolution of pattern index. Darker shaded indicate braiding and lighter shaded meandering. The black line separates the two domains. (B) Input sediment supply over time. (C) Input discharge (constant in time) as varied downstream. (D) Longitudinal profile during periods of high and low sediment supply.
Fig. 6. In- and output time series for experiment 4, in which a conceptualized warm–cold–warm scenario is being simulated. See main text for details concerning the input time series of discharge and sediment supply.

The general lines of reasoning in this scheme are:

**0 year** - An equilibrium situation (warm) exists.

**200 years** - (1) Cold sets in; evapotranspiration drops \( \rightarrow \) runoff increases; vegetation persists \( \rightarrow \) cohesion of the soil is maintained \( \rightarrow \) soil erosion (sediment supply) is low.

**300 years** - (2) Vegetation decreases \( \rightarrow \) high erodibility; high discharges \( \rightarrow \) higher sediment supply.

**400 years** - (3) Start of slow transition towards polar desert \( \rightarrow \) less precipitation; sediment supply drops slowly, but stays high.

**900 years** - (4) Start of warming; fast development of protective groundcover vegetation \( \rightarrow \) low sediment supply; high precipitation, low transpiration \( \rightarrow \) high discharges;

**1100 years** - (5) Vegetation fully developed \( \rightarrow \) high evapotranspiration \( \rightarrow \) relative low discharges and sediment supply.

Input and output time series are visualized in Fig. 6(b) and (c). Output series of elevation evolution and pattern index are shown in Fig. 6(d) and (e), respectively.

The most characteristic features of the results of this experiment are the phases of incision and aggradation, which occur both at the transition from warm to cold and from cold to warm during which phase lags exist between (changes in) sediment supply and river discharge. Morphological activity is thus linked to the *change* in climate. This behavior was predicted by Vandenberghe (1993a) and confirmed by Pleistocene river evolution (e.g. van Huissten et al., 1986; Vandenberghe, 1993b, 1994; Mol, 1997; Huisink, 1997, 1999; Tebbens et al., 1999).

The experiment further shows that channel pattern changes occur from meandering in ‘warm’ periods to braided in ‘cold’, just as classic theory predicts (Büdel, 1977). In this case, however, it is predicted by a process-response model, which makes no assumptions about the way rivers react on climate change. Note also that while incision starts during event (4) the pattern index still predicts a braided pattern. This illustrates the relative independence of channel pattern and incision/aggradation dynamics. The best explanation for the (generally) strong correlation between both is probably due to the cause of both: discharge–sediment supply ratios. The two variables, however, have different thresholds, and combinations of incision and braided may occur as well.

### 3.5. Application to the Meuse River, the Netherlands

The Meuse (Maas) is a rain fed river, that drains about 33,000 km². It originates close to the Vosges (France), crosses the Ardennes (Belgium), and flows into the Netherlands where it eventually merges with the Rhijn (Rhine) into the Rhine–Meuse delta. Mean discharge of the Meuse River is 300 m³/s. Highest
discharges occur in the winter season due to decreased evapotranspiration (Waterloopkundig Laboratorium, 1994). In this experiment, the numerical model described above is applied to the evolution of the middle reach of the Dutch Meuse during the Weichselian–Lateglacial–Holocene transition. Fig. 7 shows a summary of reconstructed climate and fluvial activity during this period.

A detailed study of longitudinal profiles of the middle reaches of the Meuse (Huisink, 1998) showed that the influences of the tectonically active Ardennes as well as sea-level changes on the evolution of the Meuse are of relatively minor significance on the time scale of the last deglaciation, and that climate is the dominant external control. It is, however, recognized that tectonics do drive the large scale evolution of the Meuse River, as shown by the river valley migration (Houtgast and van Balen, 2000), avulsions within the lower reaches (Stouthamer and Berendsen, in press) and large scale doming in the Ardennes (Garcia-Castellanos et al., 2000).

Five different terrace levels, originating from the Late Pleistocene–Holocene time interval, were recognized and analyzed (Huisink, 1998). The resulting data are shown in Table 1.

Table 1
Morphological changes in the Meuse River during the Last Glacial–Holocene transition (data from Huisink, 1998). $P$ is sinuosity, $S_t$ is terrace or floodplain gradient and $S_c$ channel gradient.

<table>
<thead>
<tr>
<th>Period</th>
<th>Age (ka BP)</th>
<th>Channel pattern</th>
<th>$P$ [-]</th>
<th>$S_t$ (cm/km)</th>
<th>$S_c$ (cm/km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Holocene (Ho)</td>
<td>&lt; 10</td>
<td>Low-sinuous meandering</td>
<td>1.3</td>
<td>12.0</td>
<td>9.2</td>
</tr>
<tr>
<td>Younger Dryas (YD)</td>
<td>11–10</td>
<td>Braided</td>
<td>1.1</td>
<td>25.1</td>
<td>22.8</td>
</tr>
<tr>
<td>Alleröd (Al)</td>
<td>12–11</td>
<td>High-sinuous meandering</td>
<td>1.5</td>
<td>23.5</td>
<td>15.7</td>
</tr>
<tr>
<td>Bölling (BÖ)</td>
<td>13–12</td>
<td>Transitional</td>
<td>1.3</td>
<td>35.1</td>
<td>27.0</td>
</tr>
<tr>
<td>Late Pleniglacial (LP)</td>
<td>&gt; 13</td>
<td>Braided</td>
<td>1.1</td>
<td>41.8</td>
<td>38.0</td>
</tr>
</tbody>
</table>
charge is therefore calculated from mean annual discharge, using a power law regression based on a database of 227 rivers worldwide for which mean annual and bankfull discharges are known (van den Berg, 1995). A least squares fit revealed that:

\[ Q_{bf} = 29.15 Q_m^{0.69} \quad (R^2 = 0.78). \]  

Bankfull discharge of the Meuse River, calculated in this way, is then \( \approx 1500 \, \text{m}^3/\text{s} \). This value is close to the current 2-year event of 1560 m\(^3\)/s (Waterloopkundig Laboratorium, 1994).

The amounts of water and sediment supply, which generated this morphology, have been assessed by the following procedure: a large number of model runs has been performed, each using temporal constant values for discharge and sediment supply. For each of these runs, resulting equilibrium channel pattern and gradient were obtained. The results of this experiment are shown in Fig. 8. Using this diagram, information on gradient and channel pattern type can be used to estimate the equilibrium sediment water fluxes.

For example, if a sand-bed palaeo channel has a gradient of 20 cm/km, and its planform type is meandering, then the stability diagram shows that bankfull discharges must have been \( < 5625 \, \text{m}^3/\text{s} \), and that transport under bankfull discharge (assuming a sand bed) has been \( < 15850 \, \text{kg}/\text{s} \).

A calculation of sediment supply based on a Holocene terrace gradient of 12.0 cm/km, a sinuosity of 1.1, a channel-forming discharge of 1500 m\(^3\)/s, and a mean discharge of 300 m\(^3\)/s, yield a channel-forming sediment supply of 780 kg/s, and sediment transport under mean flow conditions of 43 kg/s.

Extending this analysis to the past requires some extra information, namely good estimates of palaeodischarges. This issue is highly complicated due to a number of issues:

- Palaeoprecipitation during the last 20 ka is not well constrained.
- The hydrological regime was completely different during periods when mean annual temperatures were below zero and springtime

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**Fig. 8.** Stability diagram for a sand-bed river. Equilibrium channel gradient and pattern index are plotted against channel forming discharge and sediment supply.
snowmelt was a dominant runoff-generating process.

• Terrace fragments from periods when the Meuse River was braided are highly fragmented, and palaeodischarges can therefore not be made on the base of channel dimensions, flow resistance and terrace gradients.

These issues will be dealt with in detail in a subsequent paper (Bogaart, in preparation).

In the present paper, we infer bankfull discharge from AGCM experiments results (Renssen and Is-arin, in press) using the following procedure.

Mean annual temperature $T$ [°C] and total annual precipitation $P$ [mm/day] were extracted from AGCM output files for the cell representing most parts of the Meuse catchment. Total annual evapotranspiration $E$ [mm] was then calculated with the Turc equation (Jones, 1997):

$$E = P \sqrt{0.9 + \left( \frac{P}{L} \right)^2}$$

where:

$$L = 300 + 25T + 0.05T^2.$$  \hspace{1cm} (24)

The resulting rainfall excess $P - E$ [mm/day] was then converted to mean discharge $Q_m$ by summation over the whole catchment area and converting to [m$^3$/s]. A regression analysis on the data presented in van den Berg (1995) resulted in the empirical relation between mean annual flood $Q_{af}$ and mean discharge $Q_m$:

$$Q_{af} = fQ_m^{0.74}$$

where $f$ is a peakedness-factor that represents the impact of peakedness of the complete discharge time series on mean annual flood. Analysis of measured vs. predicted $Q_{af}$ for the modern Meuse River and the highly peaked Usa river (Northern Russia) results in values for $f$ of 1 and 2, respectively. Palaeo-$f$ values for the Meuse were estimated subjectively, based on inferences about the contribution of spring snowmelt to the annual discharge series.

This analysis resulted in the predictions of $Q_{af}$, as shown in Table 2.

Fig. 9 shows a stability diagram for sand-bed rivers, which is in essence a section through Fig. 8 for constant $Q$. Equilibrium gradient and channel pattern index were calculated for a range of values for sediment supplies. Palaeo sediment supply can be constrained on base of reconstructed gradients and channel pattern, and the calculated values of $Q_{af}$ in Table 2 as a proxy for $Q_{af}$. The results obtained in this way are shown in Table 3.

As can be seen from this table, are all results internally consistent, except for the Younger Dryas. A possible explanation is that the classification of the Meuse as braided during this period is false. Field inspection of the channel remnants did support this reinterpretation (C. Kasse, personal communication).

It should be noted that these results only indicate sediment transport under bankfull conditions, and thus are unlikely to represents mean transport rates. In order to obtain these latter values, a frequency magnitude analysis of all transport events must be made. This issue will also be dealt with in a subsequent paper (Bogaart, in preparation).

A tentative analysis of the modern flow frequency distribution of the Meuse River revealed that the sediment transport rates as calculated from mean annual discharge are a good approximation of the mean annual transport rates. As described above,

<table>
<thead>
<tr>
<th>Table 2</th>
</tr>
</thead>
<tbody>
<tr>
<td>Calculation of mean annual flood discharge, for five time slices of the Late Glacial, as function of estimates of climate and flow regime</td>
</tr>
<tr>
<td>Period</td>
</tr>
<tr>
<td>--------</td>
</tr>
<tr>
<td>HO</td>
</tr>
<tr>
<td>YD</td>
</tr>
<tr>
<td>Al</td>
</tr>
<tr>
<td>BÖ</td>
</tr>
<tr>
<td>LP</td>
</tr>
</tbody>
</table>
Fig. 9. Stability diagram for sand-bed rivers. The top panel shows equilibrium gradient, with annotated gradients during the Late Pleniglacial (LP), Bolling (BO), Allerød (AL), Younger Dryas (YD) and the Preboreal/Holocene (PB). The solid lines represent the cases $Q = 1500$ and $2000 \text{ m}^3/\text{s}$. The bottom panel shows equilibrium pattern index.

analysis of Holocene terraces indicate a mean sediment transport rate of 43 kg/s. This value is in accordance with values in the ‘Oxford Global Sediment Flux Database’ (Allen, 1997). The (suspended) sediment of the river Meuse as given in this database is 22 kg/s. However, van Balen et al. (2000) calculated a mean erosion rate for the last 350 ka of $\approx 8 \text{ km}^3/\text{Ma}$. This number was based on analysis of only a part of the Meuse catchment. Applying their suggested scaling ratio of 650:180 yields a total erosion rate of $29 \text{ km}^3/\text{Ma}$ which equals $29000 \text{ m}^3/\text{year}$, and, assuming a density of $2650 \text{ kg/m}^3$, a mean sediment flux of $2.44 \text{ kg/s}$. This number is a factor 20 smaller than the mean sediment transport as calculated above. Given the uncertainty within our model and its application, as well as those in the analysis of van Balen et al. (2000), it is not possible to accurately identify the source of this deviation. On one hand, our model might need further sophistication and inclusion of more processes such as reworking of in-channel sediments. On the other hand, the results of the analysis of van Balen et al. (2000) might be an underestimation: their procedure is limited to those major rivers for which terrace stratigraphies are available, and the upscaling to the whole catchment, which is now a linear function, can be improved.

Table 3
Calculated palaeo sediment discharges for the Meuse River, based on palaeogradients and channel patterns. $Q_{s,g}$ is sediment discharge [kg/s], based on gradient. $Q_{s,p}$ is sediment discharge, based on channel pattern.

<table>
<thead>
<tr>
<th>Period</th>
<th>Gradient</th>
<th>Pattern</th>
<th>$Q_{s,g}$</th>
<th>$Q_{s,p}$</th>
<th>Consistency</th>
</tr>
</thead>
<tbody>
<tr>
<td>Preboreal</td>
<td>12</td>
<td>M</td>
<td>$\approx 900$</td>
<td>$&lt; 10800$</td>
<td>yes</td>
</tr>
<tr>
<td>Younger</td>
<td>25</td>
<td>B</td>
<td>$\approx 11500$</td>
<td>$&gt; 14100$</td>
<td>no</td>
</tr>
<tr>
<td>Dryas</td>
<td>23</td>
<td>M</td>
<td>$\approx 6800$</td>
<td>$&lt; 12900$</td>
<td>yes</td>
</tr>
<tr>
<td>Allerød</td>
<td>35</td>
<td>B/M</td>
<td>$\approx 11200$</td>
<td>$\approx 11700$</td>
<td>yes</td>
</tr>
<tr>
<td>Bolling</td>
<td>42</td>
<td>B</td>
<td>$\approx 22300$</td>
<td>$&gt; 12600$</td>
<td>yes</td>
</tr>
<tr>
<td>Late</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Pleniglacial</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
4. Discussion and conclusions

A numerical model is presented that, although very simplified, is founded on a process-based paradigm. The model operates on the Quaternary time scales of 0.1–100 ka.

The results of the first experiment with this model show that phases of incision and aggradation can be explained by changes in sediment supply or discharge, which are not necessarily coupled. It also shows that, in this case, there is a correlation between incision/aggradation and channel pattern. One cannot, however, conclude directly that channel pattern is controlled by incision/aggradation only. Instead, it is proposed that both incision/aggradation and channel pattern are a function of the ratio between sediment supply and bankfull discharge.

The results of the second experiment show that the time lag between changes in river discharge and sediment supply determine the river response to these changes, although further field evidence is required to test this model. These results have major implications for the reconstruction of climate events based on documented river response in the geological record. They show that incision/aggradation is not necessarily coupled to (climate induced) changes in water and sediment supply as such, but more precisely to the phase-lags between them. Therefore, one cannot use observed phases of incision and aggradation to infer changes in sediment supply without considering discharges as well, for example, because the effect of discharge changes is unknown. However, the link between reconstructed incision/aggradation events and palaeohydrology can still be made, for example by taking palaeo channel width into account. Since channel width is determined by (bankfull) discharge directly, it can be used as a proxy for the latter. In a hypothetical situation where no net aggradation or incision occurred over a series of known climate transitions, while changes in channel width, hence discharge, did occur, the conclusion must be that sediment supply changed as well, in phase with discharge.

The results of the third experiment show that a gradual downstream, temporarily varying increase in discharge, can result in a complex response of channel pattern, which varies for different sub-reaches of the river. The implication of this result is that river pattern is not only dynamic in time but also in space, because downstream variations in the controlling factors (see also Tebbens, 1999; Veldkamp and van Dijke, in press) are not. It is part of the canonical fluvial geomorphological knowledge that rivers tend to braid in their upper reaches, and meander in their lower reaches (and even change to anastomosing in their deltas). Therefore, the modeling results do not extend current knowledge per se, but they do provide a quantitative explanation for field-derived knowledge, and are a tool to assess the relative importance of the different driving forces behind river response to environmental change.

In experiments 4 and 5, it is further shown that our model is also capable of simulating climate transitions, both for conceptual and actual rivers. Although validation against reconstructed climate data and fluvial evolution remains necessary, the preliminary model applications reported here already show agreement with these reconstructions in both a qualitative and quantitative way.

The model can easily be criticized on the underlying assumptions and choices, as “[this class of models] has assumptions that are false and that are known to be false” (Beven, 1997). However, this should not hamper application of the model to real-world problems, provided the model is not used as a prediction tool, but as a conceptual/theoretical connection between input data and river response. More often than not, one might see that the model does not predict very well the actual quantitative morpho-response for a given time series of input.

The origin of this error might lie in the model structure, in the input data, or in the data to which the model output is compared. Especially in palaeo-research, these data are the results of indirect observation like proxy records or inference, and thus more uncertain than desired. For this reason, new insights can be acquired by comparing model results with observations, and by correlation of the data-model deviations with the three potential error sources outlined above.

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